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Evaluation of Nimrod radar corrections - Report to the Nimrod project board

by

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Note

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1 Introduction

An automated version of the radar analysis employed by the FRONTIERS system is being developed for use in Nimrod. This report describes the evaluation of the new method of correcting the radar rainfall rates to produce the best estimate of the surface value. The automation of the deletion of spurious echoes is being undertaken by the Nowcasting Development Group and is not considered here.

Two methods are used currently within the Meteorological Office to derive surface precipitation rate estimates from weather radar data. At the radar sites, fixed corrections for occultation and range are applied and in addition, at about half the 13 radars in the network, gauge adjustment is performed using the scheme of Collier et al (1983). The radar site corrections are removed within FRONTIERS and more sophisticated corrections applied (see Brown et al, 1991). The FRONTIERS corrections are potentially more accurate than the site corrections, but involve human decisions and the output is not as timely; being available some 15 minutes after the data collection time.

Evaluations of the new method have been performed using two types of data as ground truth. Initial evaluations used high resolution reflectivity profiles from the Chilbolton experimental radar. Using this data, it was possible to compare instantaneous values of reflectivity and these comparisons are the most valid scientifically. Because simultaneous Chilbolton and Wardon Hill data were only available for February and March 1991, nearly all the cases studied were frontal and the average rainfall rate was generally less than 1 mm/h. These cases were not the most challenging for the new scheme, but neither did they demonstrate its full potential.

A further series of cases were then chosen encompassing a wide range of synoptic types. Some of these were very challenging, others contained severe errors allowing the full potential to be demonstrated. Chilbolton data were unavailable for these cases, necessitating the use of rain gauges as ground truth. There are several disadvantages to the use of gauges. It is necessary to integrate the radar data to produce hourly or daily totals which can lead to temporal representativeness errors because the gauges sample continously whilst the radar data comprise snapshots at 5 or 15 minute intervals. There is also a spatial representativeness error because the gauge samples at a point whilst the radar data are 5x5km averages. Finally, the reflectivities have to be converted to rainfall rates using an empirical Z-R relationship. The net result is that a larger scatter is expected in the results using gauges, compared to the use of Chilbolton profiles, which may partially mask the true benefits of the new algorithms.

The Nimrod Stage 2 Plan requires that the automated radar analysis be as good as FRONTIERS on 95% of occasions. However, FRONTIERS quality-controlled data is only available at 15-minute intervals and would produce significant temporal errors in comparisons with gauges (Kitchen and Blackall, 1992). The quality-controlled data is also only archived as a composite field, whilst the new radar correction scheme is designed to work with single-site data. Therefore, the quantitative evaluations were performed using remote site data from individual radars because this was available at 5-minute intervals. Although the FRONTIERS quality-controlled data were not evaluated quantitatively, previous evaluations found no significant difference between FRONTIERS and remote-site data (see Warner, 1990). Thus, if it can be demonstrated that after application of the

new algorithms, the radar data are significantly more accurate than remote site data, it can safely be assumed they are at least as good as FRONTIERS quality-controlled data. To check on this, the FRONTIERS fields were compared subjectively with the results of the new technique.

2 The New Correction Method.

2.1 Outline of the method.

Recent publications on errors in estimating surface rainfall from weather radar data have suggested that given a correctly calibrated radar, the dominant source of error is due to the radar beam not sampling the precipitation close enough to the ground (Joss and Waldwogel, 1990, Andrieu and Creutin, 1991). The FRONTIERS radar analysis attempts to allow for this using corrections based upon 6 prescribed reflectivity profiles. Kitchen and Jackson (1993) showed that significant variability (in terms of correcting radar data) in the vertical profile of reflectivity occured on the scale of individual pixels. They concluded that for a correction method to have the greatest potential benefit, it should be capable of correcting on a pixel-by-pixel basis (see also Smith, 1990). The reasons for the failure of the FRONTIERS corrections are that only 6 profiles are available to deal with the infinite variety of profiles which actually occur and the same profile is used over a large area.

The key feature of the Nimrod correction method is that a different idealized profile of reflectivity is constructed in real time *at each pixel.* A typical profile is shown in Figure 1. The reflectivity increases downwards from the precipitation top to the freezing level. Beneath the freezing level is located a bright-band peak and beneath this the reflectivity is either constant to the surface or increases linearly beneath 1.5km when orographic enhancement is diagnosed as occuring. The region of constant reflectivity beneath the bright-band peak is known as the background reflectivity and the correction method seeks to calculate this. The known beam power profile of the operational radars is used to compute the measured reflectivity expected given an initial guess of the background reflectivity. This is compared with that actually observed. The background reflectivity is then altered and the calculation repeated. The iteration proceeds until the calculated and observed reflectivity agree to within 10%. Iteration is required because the area under the bright band is a non-linear function of the underlying reflectivity. The background reflectivity is converted to the equivalent precipitation rate using the standard Z-R relationship applied in FRONTIERS. Finally, the surface precipitation rate is estimated by adding any orographic enhancement that is diagnosed onto the background rate. The variable parameters defining the idealized profile, eg precipitation-top height, are diagnosed *at each pixel* in real-time using relevant meteorological observations and model forecast fields. Even if current observations and understanding do not permit the smaller scale variations in the profile to be resolved, such a scheme has the potential for improvement in the future.

2.2 Specification of the reflectivity profile.

The idealized reflectivity profile was derived from an initial study using RHI scan data from the S-band Chilbolton radar operated by the Rutherford Appleton Laboratory. Scans

from a total of 112 days over a 3 year period were analysed and vertical profiles constructed. The profiles have a resolution of 200m in the vertical and are averages over a 5.3km line in the horizontal. The profiles were used to examine the relationship between the intensity of the bright band and the underlying rainfall rate, the shape of the profile above the bright band, and orographic growth at low levels. The results were then combined to form the idealized reflectivity profile shape shown in Figure 1 which is defined by a small number of parameters. These are;

- the background reflectivity;
- the height of the freezing level;
- the height of the top of the precipitation layer;
- the magnitude of the anticipated orographic enhancement.

In addition, other fixed parameters required at each pixel are;

- the ground height above sea level;
- the radar beam elevation angle;
- the radar range;
- the radar beam occultation angle.

2.3 Requirement for satellite, synoptic and mesoscale model data.

Two alternative methods may be used to diagnose the height of the freezing level, the depth of the precipitation, and to specify the orographic correction field. The simplest methods involve the use of SYNOP observations and METEOSAT infrared data only and these are described first.

Screen temperature and dew point observations are used to calculate the wet bulb freezing level at each observation site. A recursive filter analysis (Purser and McQuigg, 1982) is then applied to provide a field of values at each pixel. The reliance on surface observations can lead to significant errors in the wet bulb freezing level where precipitation is falling through stable layers such as occur close to warm fronts. However, the effect of errors in height upon the corrections are mitigated by the assumption of a relatively deep melting layer and the correction method has so far proved very successful at reducing errors due to the bright band.

Hill (1983) specified the orographic enhancement in terms of the wind speed and direction at 800m altitude. Subsequently, he also introduced a humidity scaling factor to reflect the moisture supply into the feeder cloud. In the adaptation to utilise surface observations, the observed 10m wind components are analysed to provide an estimate of the 10m wind speed and direction in each pixel. The 800m wind speed and direction is then estimated using a climatalogical correction. There is currently no humidity scaling save the imposition of a humidity threshold. If the analysed surface relative humidity is below the threshold value (currently set at 85%), the orographic correction is set to zero.

Precipitation top height is difficult to define from available observations. In strong convection, precipitation-size hydrometeors may extend right up to cloud top whereas in

frontal cloud the cloud depth may be very much larger than the depth of the precipitation layer. Currently, the method uses infra-red images from METEOSAT combined with radiosonde soundings (if available) to convert the cloud-top temperature to an equivalent height. This height is then assumed to be the top of the precipitation layer except in two circumstances. If the top of the precipitation is diagnosed to be at a height of less than 1.5 km, then significant precipitation is considered highly unlikely and 1.5 km is adopted as the minimum. A situation which typically occurs in wintertime warm fronts is where the freezing level is very low with very cold cirrus cloud extending well above the main frontal precipitating cloud layers. In such conditions, precipitation development is not associated with the observed cloud top, but much closer to the freezing level. Some correction for range is normally required in these conditions and to ensure that some is applied, the precipitation top in the idealized profile is limited to being no more than 4km above the freezing level.

Alternatively, forecast temperature fields from the most recent run of the mesoscale model may be used to derive a freezing level field. During the period encompassing the evaluation case studies, the version of the mesoscale model in operational use was that described in Golding (1990). At each model grid point, the freezing level is found by interpolation between model levels. Where multiple freezing levels are indicated due to the presence of inversions, the lowest level is selected. Finally, the height at each radar pixel is interpolated from the grid point values. The forecast windfield and relative humidity at the 810m level is used to select the orographic correction field at each grid point and to apply a humidity scaling factor. If the relative humidity is below 85%, then no corrections are applied and the scaling factor increases linearly up to a maximum of 1.5 at 100% (Davies, 1993). Again, interpolation is required to derive radar pixel values from the model grid.

For those grid boxes where the model produces dynamical rainfall, the height of the top of the precipitating layer may be estimated from the fields of cloud water (or ice) content. Heymsfield (1977) derived an empirical relationship between ice water content and radar reflectivity from simultaneous aircraft measurements of particle size distributions and ground-based radar measurements.

$$Z = 750.I^{1.98} \tag{1}$$

where Z is the reflectivity in mm^6m^{-3} and I is the ice water content in gm^{-3}. Using this relationship, the top of the precipitating layer was defined as the first model level (from the top level downwards) for which the estimated reflectivity exceeded 1 mm^6m^{-3}. Where model rainfall is produced by the sub-grid scale convective parametrization, the precipitation top is assumed to be at the cloud top height. In those grid boxes where both types of precipitation are indicated, the type associated with the lower surface precipitation rate is ignored for the purpose of estimating the precipitation depth. For pixels where the model forecast surface precipitation rate is zero, the precipitation depths estimated directly from METEOSAT IR data are substituted.

2.4 Calculation of the corrections.

As the intensity of the bright band is not directly proportional to the background reflectivity, an iterative method is required to derive a surface rainfall rate estimate from the measured reflectivity. An intermediate value of the background reflectivity is first selected (currently 200 mm^6m^{-1}) and the reflectivity profile constructed. The beam power profile is convoluted with the reflectivity profile to calculate what reflectivity the radar would measure ($\overline{Z}$) using the following equations (from Brown et al., 1991).

$$\overline{Z} = \int_{\alpha}^{\beta} Z(\phi) f(\phi) d\phi \tag{2}$$

where ϕ is the elevation angle, $f(\phi)$ is the fraction of the radar beam power in the range ϕ to $\phi + d\phi$ and the beam power is negligible outside the range of angles α to β.

$$f(\phi)d\phi = [P(\phi)d\phi] / \int_{\alpha}^{\beta} P(\phi)d\phi \tag{3}$$

and $P(\phi)d\phi$ is the relative power returned from angles in the range ϕ to $\phi + d\phi$. The transmitted beam power profile relative to the beam center ($\phi = 0$) specified by the radar manufacturer is well fitted by the function $[\sin(k\phi)/k\phi]^2$ (Brown, 1987), where k is a constant (=159.46 for ϕ in degrees). The function falls to zero for $|\phi| > 1.13^o$ and the power is assumed to be zero outside this range. The half-power beam width is 1^o. Therefore the received power profile is given by:-

$$P(\phi) = [\sin(k\phi)/k\phi]^4 \tag{4}$$

$\overline{Z}$ is found by integration of equation 1 using 0.03^o angular increments. α is the radar horizon at a particular azimuth.

In general, the calculated reflectivity will not equal that which was measured and so the integration is repeated using different values of the background reflectivity until a value is found which when used to define the profile, results in agreement between calculated and measured reflectivities to within some tolerance. The value of the fractional tolerance adopted is 0.1. Using the bisector method, convergence is normally achieved within about 10 iterations. The reflectivity at ground level associated with this background reflectivity is then translated into rainfall rate using the assumed $Z - R$ relationship ($Z = 200R^{1.6}$).

Colour plate 1 shows an example of a corrected Dyfed radar image and the site corrected data from which it was derived. In this case, the freezing level was below 1km height and the remote site data has a bright band in both infilling beam and surface beam areas. The correction scheme largely eliminates this spurious structure. At longer ranges, the corrections increase the precipitation rates relative to remote site. Qualitatively, the corrections appear to have been beneficial in this case. Quantitative evaluation of these data is provided in the next section.

It is intended that, in operational use, the correction method will be applied to data from each radar separately. A composite image of the data from the network will then be formed from these corrected data.

3 Evaluation of the Corrections.

3.1 Comparison methods.

The preliminary evaluation of the correction method was performed by comparisons between corrected Wardon Hill radar data and estimates of surface precipitation rate derived from near-surface reflectivity measurements by the Chilbolton radar during February and March 1991 (see Kitchen et al, 1993). During this period, the Chilbolton radar executed series of RHI scans along 24 regularly spaced azimuth angles and data from ranges up to about 60km were considered for comparison purposes. For 7 wet days during the two months, the data from the Wardon Hill operational radar were corrected using the new method in a realistic simulation of operational conditions (see Table 1). Data from Chilbolton RHI scans within each pixel were grouped into height bands 200m deep. The 5km pixel data from Wardon Hill were then compared with collocated Chilbolton reflectivity factors for the layer centred at 500m. The 500m reflectivity factor was taken to be the best estimate of the true surface precipitation rate. All reflectivity factor profiles were subject to quality control to ensure that there was adequate data from the 500m level and to minimise contamination by ground clutter or occultation. Comparisons were also rejected from within a range of azimuths from Wardon Hill, within which the Wardon Hill radar suffers from severe occultation. The comparison pixels were at ranges between 40 and 125 km from Wardon Hill, i.e. at ranges typical of the operational network coverage of the UK land area. Data were considered to be collocated if both radars sampled the same pixel within two minutes of each other. In two minutes, the spatial pattern of rainfall may be expected to be displaced by no more than half a pixel width at typical advection speeds.

The use of line average reflectivity factors as an estimate of areal averages over the 5km pixels incurs representativeness errors (ϵ). These were quantified by comparing line averages in a number of pixels close to the Chilbolton radar which were intersected by RHI scans along two or three adjacent azimuths.

As noted in the introduction, the Chilbolton comparisons did not constitute a sufficient test of the method because of the limited range of synoptic conditions. To try and demonstrate the efficacy of the corrections over a wide variety of conditions, 6 case studies were performed in which comparisons were made between gauge accumulations and radar integrations. All the studies were from the the period November 1991 - November 1992 and are listed in Table 3. The cases were selected without reference to, or knowledge of, radar measurement errors. The method was developed initially using analyses of freezing level etc. based directly upon observations and one set of evaluations used these analyses only. Since it was hoped that mesoscale model forecasts would provide superior analyses for use in Nimrod, another set of evaluations was done with mesoscale model data introduced into the diagnoses of freezing level, orographic corrections and precipitation top. Portions of each case were therefore repeated using corrections derived using model data. Time did not allow an assessment of the impact of the model data when used to determine only one or two of the three fields.

The corrections are intended to be applied to data from individual radars rather than a network composite. Comparisons were therefore between gauge accumulations and integrations of collocated data from a single radar. Data from both the Met. Office hourly

and daily gauge networks were used as appropriate. All gauges out to the maximum radar range of 210km were considered. Note that when the network radar data is composited, the typical maximum range for radar coverage over England and Wales is about 150km.

In previous tests covering periods in 1990 and 1991, Wardon Hill radar had been found to have a detection limit consistent with the manufacturers specification (Kitchen and Jackson, 1993) and to be correctly calibrated (Kitchen et al, 1993). Comparisons with the Chilbolton radar had also enabled the beam elevation angle to be independently determined. Given this history, it had been intended to make use of Wardon Hill data in the present comparisons, along with Chenies and Dyfed. However, systematic bias errors were found in one of the cases which the current version of the radar correction scheme cannot compensate for and consequently Wardon Hill data were excluded from the gauge comparisons.

The aim of the corrections is to produce a field of surface precipitation rate. If the radar sometimes fails to detect precipitation, radar integrations may be lower than collocated gauge accumulations although the non-zero instantaneous rates may be perfectly accurate (see Kitchen and Jackson, 1993). To reduce the impact of these errors, hourly comparisons were excluded where one or more radar measurements during the hour were zero, except in cases of scattered convection where the data loss incurred would have been almost total. Daily gauge comparisons with less than 12 hours of collocated radar precipitation were similarly excluded.

3.2 Quantitative comparisons

The comparison statistics for each of the Chilbolton comparison cases are presented in Table 2. In the table, N is the number of satisfactory collocations, S denotes the surface precipitation rate estimate derived from Chilbolton data, R is the raw Wardon Hill precipitation rate and C is the corrected surface precipitation rate estimate from the Wardon Hill data.

The best estimate of the true errors in the surface precipitation rate estimates were computed by subtracting ε^2 from the mean square comparison differences and the results are shown in the last three columns of Table 2. In most cases, the impact of representativeness errors upon the error estimates was small; confirming one advantage of verification against radar rather than gauge data. In the last two cases in Table 2, however, the residual RMS errors were of comparable magnitude to ε. In these cases, the RMS error estimates are more uncertain, particularly in view of the non-normal difference distributions. The final column in Table 2 shows the fractional reduction (expressed as a percentage) in RMS error achieved using the new corrections compared to remote site data. The correction method consistently improved the surface precipitation rate estimates with fractional reductions in RMS errors in the range 10-78%. The mean differences, whether overestimation due to bright band or underestimation due to the effects of orographic growth or beam overshooting, were also reduced in almost all cases. Figs. 2 and 3 are examples of scatter plots from two cases showing large reductions in error.

The gauge comparison results are summarized in Table 4. In the table, N is the number of individual comparisons (hourly or daily); G denotes accumulations measured by gauges; R denotes integrations of remote site radar data and C denotes integrations of

corrected radar data with the corrections derived without using model data. The figure in the final column is the fractional change (RMS (G-R) - RMS (G-C)) / RMS (G-R) expressed as a percentage. Therefore positive values of the fractional change indicate that the new scheme reduces RMS compared to remote site. In 8 of the 9 periods, the Nimrod corrections reduced RMS differences compared to remote site. In table 5, the performance of corrections with and without the benefit of model data are compared. M denotes integrations of corrected radar data produced using the model derived fields. The last two columns enable the fractional reduction in differences to be compared. The introduction of model data improved the corrections in the first case but produced corrections which appear to have been significantly worse than the observation based corrections in at least two other cases.

3.3 Qualitative evaluation of the gauge comparison cases.

a) 13 Nov 1991

The correlation between radar and gauge accumulations was poor (see Fig 1) and typical of highly convective conditions where the spatial variability in precipitation rate may be concentrated on the sub-pixel scale. At long range, the corrections applied to the data resulted in large overestimation of some gauge accumulations, and this dominated the overall statistics. Two possible causes are problems associated with radar measurements of precipitation in this case. Firstly, many SYNOP reports in the area covered by Dyfed radar were of hail showers. The standard Z-R relationship used in the corrections will tend to overestimate precipitation rates for hail. Also, the bases of the Cb clouds producing the showers were reported to be at heights of typically 500-600m above the ground whereas the wet bulb freezing level was estimated to be at about 500m. Thus there is the possibility that snow falling from the clouds may have suffered evaporation before melting. The freezing level heights derived from the model were slightly higher than those derived from the SYNOP observations. Similarly, the precipitation tops in the model were more than 4km above the freezing level. These factors combined to produce smaller corrections at longer ranges when using model data; with the result that the model corrections were overall closer to remote site (Table 5).

In FRONTIERS, range corrections were applied which reduced intensities at long range compared to remote site. Thus it seems probable that FRONTIERS corrections were superior to the Nimrod corrections in this case.

b) 8/9 Jan 1992

There was evidence of significant temporal changes in the shape of the reflectivity profile above the melting layer in this case. Accordingly, the hourly gauge comparisons were split into three periods. During the period 1800-2400UTC on the 8 Jan, at ranges where the radar beam was largely above the melting layer, the corrections were too small and a pronounced range effect was evident in the corrected radar fields. The corrections performed best during the period 0000-0400UTC on the 9 Jan when both the bright band and range errors were largely mitigated (Fig 4). Conversely, at times after 0400UTC on the 9 Jan, the gauge comparisons suggested that corrections at long range resulted in

overestimation. Almost all the precipitation was associated with very deep cloud and the condition that the diagnosed precipitation top height should be no more than 4km above the freezing level was widely applied. Despite these difficulties, the corrections reduced the radar/gauge differences in all three periods. The results from the radar/daily gauge comparison suggested a reduction in RMS error of 47% overall. These results, and others not presented here, show that there is scope for improvement in the diagnosis of reflectivity profile shape above the melting layer.

The new scheme was clearly superior to FRONTIERS because no bright-band correction was applied in FRONTIERS, despite clear visual evidence of bright-band contamination in at least one image. The FRONTIERS range corrections were either similar or slightly smaller than those applied in the new scheme.

c) 9 Jun 1992

As in the 13th Nov case, the precipitation was from thunderstorms with some hail. The correlation between radar integrations and gauge accumulations was small. However, there is evidence that the Chenies radar also systematically overestimated gauge accumulations. The reason is unclear, but the known presence of hail and some anaprop could be contributory effects. A radar calibration error cannot be ruled out. The freezing level over south east England was above 3km and so the corrections applied to the radar data were relatively small. Overall, the corrections resulted in a marginal reduction in the RMS difference (Table 4).

Comparison with FRONTIERS was difficult because of the small scale of the precipitation features and also because there was no data at long range from Chenies in the FRONTIERS composite. No systematic differences between FRONTIERS and the new scheme were apparent.

d) 3 Jul 1992

A summertime case of very widespread frontal rainfall with embedded convective cells producing heavy rainfall. Correction of bright band overestimation at long range resulted in a reduction of RMS radar/gauge differences by 50% overall (Fig 5).

There was no evidence of bright band correction in FRONTIERS but some benefit will have derived from removal of the remote site range correction and substitution of smaller corrections, thereby reducing intensities at long range. In this case, it is difficult to say whether Nimrod corrections represent a significant improvement over FRONTIERS but they were clearly as good.

e) 20/21 Jul 1992

An intense bright band associated with heavy rainfall led to very large errors at long range in the remote site radar data. The maximum hourly accumulation recorded by the gauges was about 30 mm whereas the raw radar data persistently indicated rates of >64 mmh^{-1} over large areas at longer ranges. The bright band errors were reduced significantly but not eliminated by the corrections (Fig 6). The RMS differences were

reduced by nearly 80%, but remained rather high at $5.93mmh^{-1}$ compared to a mean gauge accumulation of $5.41mmh^{-1}$. The mean error was reduced by about 70%.

From Table 3, the mesoscale model fields proved slightly inferior relative to those derived from SYNOP and satellite data directly.

In the FRONTIERS composite, no bright band correction was applied and the precipitation rates at long range were similar to remote site. There is no doubt that in this case, the Nimrod corrections represented a large improvement over FRONTIERS.

f) 29/30 Nov 1992

The spatial distribution of rainfall was dominated by orography. Frontal rainfall intensified over Wales with Dyfed remote site rates of $> 8mmh^{-1}$ over the S. Wales valleys and Snowdonia. The freezing level was at about 2km and the presence of the bright band centered at a range of about 100km resulted in a general reduction in the highest precipitation intensities over the mountains, despite orographic corrections being prescribed. This surprising result is nevertheless supported by the gauge comparison data. At ranges $> 75km$, the radar integrations overestimated remote site gauge accumulations and the corrections achieved a reduction in RMS errors of 62%. At shorter ranges, corrections were small and over all ranges, the reduction in RMS errors was 32%.

The Nimrod scheme applied orographic corrections which appeared similar to those applied in FRONTIERS. It is impossible to compare the range corrections because there is no data at long range from Dyfed in the FRONTIERS composite.

4 Conclusions.

The comparisons with Chilbolton radar data showed that the new corrections reduced RMS errors in surface precipitation rate by between 10 and 78% compared to the remote site radar corrections.

Application of Nimrod corrections to a wide range of synoptic types resulted in a reduction of RMS differences between radar and gauges compared to remote site in all except one case. Most of the improvement originated from correction of overestimation due to the bright band. The results of radar/gauge comparisons are difficult to interpret, especially in convective precipitation where representativeness errors are large. Therefore it should be noted that the RMS differences in Table 4 are not the RMS errors in the radar estimated surface rainfall accumulations which may be significantly smaller.

From the results presented in Table 4 it appears that the Nimrod corrections are superior to those applied at the radar sites. Since previous evaluations showed no significant difference between the performance of the remote site and FRONTIERS corrections eg Warner (1990), it can be inferred that the new scheme is superior to the FRONTIERS scheme. Subjective comparisons of the corrected fields from the Nimrod scheme and from FRONTIERS showed that the Nimrod scheme was comparable or better than the FRONTIERS scheme. It was certainly more consistent than FRONTIERS at correcting bright-band errors; even serious cases sometimes remaining uncorrected in FRONTIERS.

The use of mesoscale model data in the diagnosis of reflectivity profile shape resulted in significantly better corrections in only one of the cases. It is believed that the precipitation top height is the field most likely to have been responsible for the negative impact of the model analyses in other cases. Unfortunately, it was not possible to test the impact of model data on freezing level, orographic corrections or precipitation top separately.

It should be emphasized that the corrections rely upon accurately calibrated radars and some data were rejected from the comparisons for this reason. The corrections will be compromised if correct radar sensitivity cannot be maintained across the network. Also, the corrections are sensitive to errors in the radar elevation angle of order 0.1 deg. There is presently uncertainty of of the same order concerning the true beam elevation angle data as supplied by the radar manufacturers.

5 Recommendations.

- The new correction procedure should be implemented in Nimrod.
- Action is necessary to ensure that the quality of the precipitation analyses are not compromised by variations in radar sensitivity or uncertainty in radar beam elevation angle. As a last resort, some form of gauge adjustment will have to be introduced into the scheme.
- There are areas where the new correction scheme can be significantly improved. Development should continue outside the formal Nimrod project during the time leading up to operational implementation.
- The precipitation top field should continue to be produced from METEOSAT IR data directly. The facility for deriving freezing level and orographic corrections from SYNOP data should be retained as an option or backup until such time that it can be demonstrated that model derived fields provide improved corrections.

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Table 1

Chilbolton radar evaluation cases.		
Date (1991)	Time period (UTC)	Synoptic conditions
7 Feb	0500-1000	snow showers
21 Feb	1500-1800	warm sector
22 Feb	1400-2000	warm front/sector
27 Feb	1600-1900	weak fronts
8 Mar	0000-0500	warm front/sector
16 Mar	1100-1530	cold front
18 Mar	1200-2300	warm front

Table 2

Chilbolton radar comparison results.										
		mean		mean	mean	RMS	RMS	RMS error estimates		
Date (1991)	N	S (mm/h)	ε (mm/h)	S-R (mm/h)	S-C (mm/h)	S-R (mm/h)	S-C (mm/h)	R (mm/h)	C (mm/h)	reduc (%)
7 Feb	15	0.27	0.02	-0.17	-0.12	0.28	0.25	0.28	0.25	13
21 Feb	63	0.70	0.16	0.47	0.09	0.85	0.32	0.83	0.28	67
22 Feb	136	0.94	0.18	0.18	-0.10	0.92	0.47	0.91	0.44	52
27 Feb	43	0.56	0.06	0.10	-0.08	0.29	0.19	0.28	0.18	36
8 Mar	95	0.57	0.19	0.17	-0.05	0.54	0.28	0.51	0.21	58
16 Mar	84	1.04	0.52	0.48	0.00	1.49	0.60	1.39	0.30	78
18 Mar	148	0.46	0.17	-0.07	-0.07	0.24	0.23	0.17	0.16	10

Table 3.

Gauge evaluation case studies.		
Date	Radar	Synoptic Conditions
13^{th} Nov 1991	Dyfed	scattered convection with hail
8^{th}-9^{th} Jan 1992	Dyfed	widespread precipitation circulating depression
9^{th} Jun 1992	London	thunderstorms with hail
3^{rd} Jul 1992	London	widespread mainly convective rainfall
20^{th}-21^{st} Jul 1992	London	multiple intense mesoscale convective storms
29^{th}-30^{th} Nov 1992	Dyfed	frontal with orographic rainfall

Table 4.

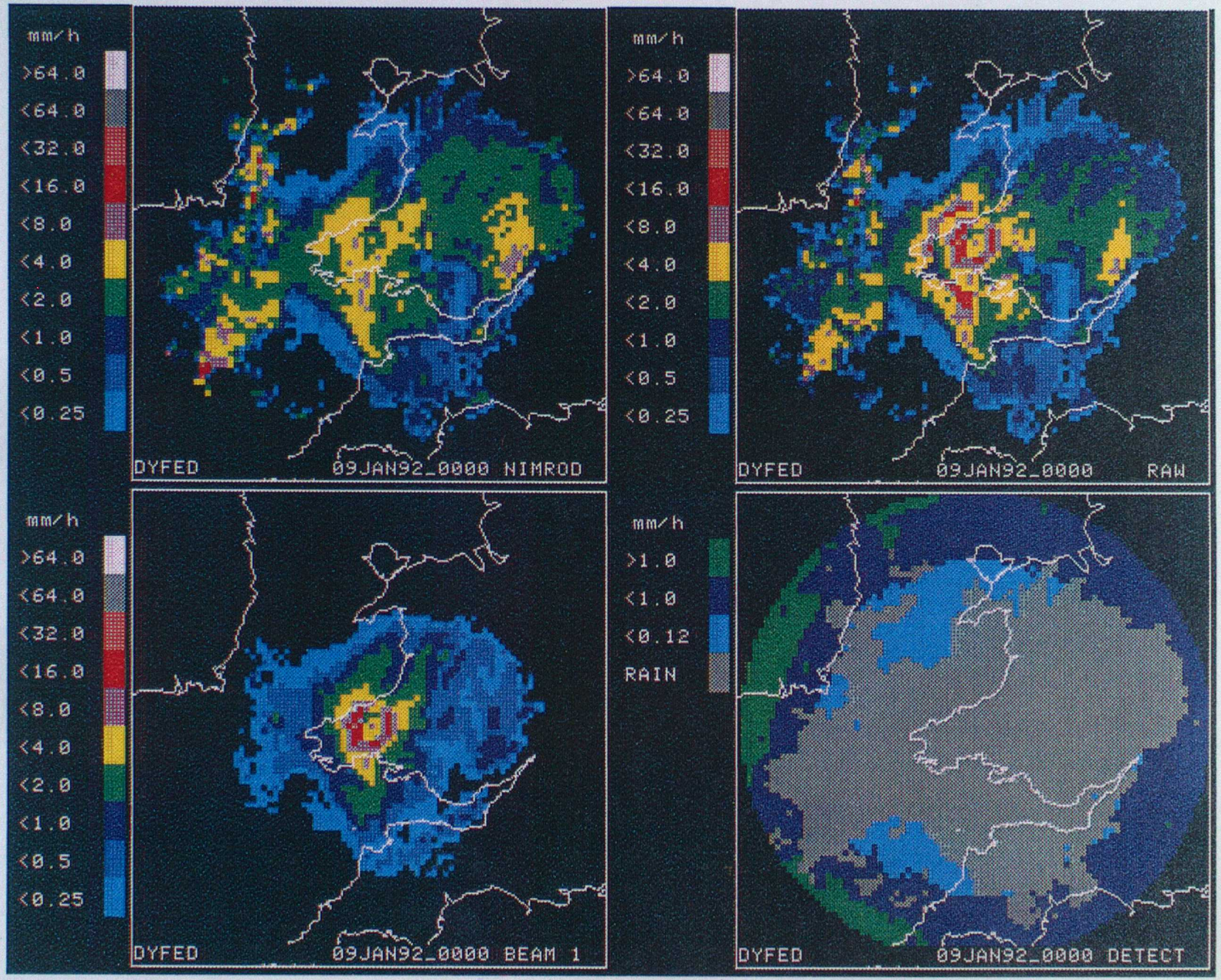
Results of radar-gauge comparisons (no mesoscale model data used)									
Date	Time Period (UTC)	Hrly/ Daily	N	mean G (mm)	mean G-R (mm)	mean G-C (mm)	RMS G-R (mm)	RMS G-C (mm)	reduc G-C %
13 Nov 1991	0000-1200	H	84	0.59	0.12	-0.22	0.55	0.83	-51
8 Jan 1992	1800-2400	H	57	1.69	0.34	0.60	1.76	1.40	20
9 Jan 1992	0000-0400	H	34	1.74	-0.55	-0.02	2.10	0.74	65
9 Jan 1992	0400-0900	H	47	3.06	-2.01	-1.51	6.73	3.71	45
8/9 Jan 1992	0900-0900	D	386	31.18	-0.07	2.73	24.52	13.11	47
9 Jun 1992	1300-1700	H	26	4.34	-12.46	-7.74	22.01	13.64	38
3 Jul 1992	1200-1800	H	72	2.25	-0.88	-0.03	2.11	1.06	50
20/21 Jul 1992	1900-0400	H	69	5.41	-13.32	-3.35	25.48	5.93	77
29/30 Nov 1992	0900-0900	D	169	29.06	-10.73	-4.42	17.61	12.02	32

Table 5.

Radar-gauge differences with and without the utilisation of mesoscale model data.									
Date	Time Period (UTC)	Hrly/ Daily	N	mean G (mm)	RMS G-R (mm)	RMS G-C (mm)	RMS G-M (mm)	reduc G-C %	reduc G-M %
13 Nov 1991	0000-1100	H	84	0.57	0.55	0.83	0.63	-51	-15
8 Jan 1992	1800-2400	H	57	1.69	1.76	1.40	1.37	20	22
9 Jan 1992	0000-0400	H	34	1.74	2.10	0.74	1.05	65	50
9 Jan 1992	0400-0900	H	47	3.06	6.73	3.71	7.27	45	-8
8/9 Jan 1992	0900-0900	D	386	31.18	24.52	13.11	15.27	47	38
9 Jun 1992	1300-1700	H	26	4.34	22.01	13.64	17.66	38	20
3 Jul 1992	1200-1800	H	70	2.25	2.11	1.06	1.05	50	50
20 Jul 1992	1900-2200	H	37	6.11	23.30	5.24	6.45	78	72
29/30 Nov 1992	0900-0900	D	169	29.06	17.61	12.02	16.81	32	5

Plate 1

Example of typical output from the correction scheme.
Top LHS - corrected image from Dyfed radar for 0000UTC 9 Jan 1992.
Top RHS - remote site image for the same time showing the spurious double bright band structure largely eliminated by the corrections.
Bottom LHS - beam 1 data which is not used in the current version of the correction scheme.
Bottom RHS - diagnostic output which is used in the merging of precipitation derived from satellite data with the radar data.


mm/h
>64.0
<64.0
<32.0
<16.0
<8.0
<4.0
<2.0
<1.0
<0.5
<0.25
DYFED
09JAN92_0000 NIMROD
mm/h
>64.0
<64.0
<32.0
<16.0
<8.0
<4.0
<2.0
<1.0
<0.5
<0.25
DYFED
09JAN92_0000 RAW
mm/h
>64.0
<64.0
<32.0
<16.0
<8.0
<4.0
<2.0
<1.0
<0.5
<0.25
DYFED
09JAN92_0000 BEAM 1
mm/h
>1.0
<1.0
<0.12
RAIN
DYFED
09JAN92_0000 DETECT

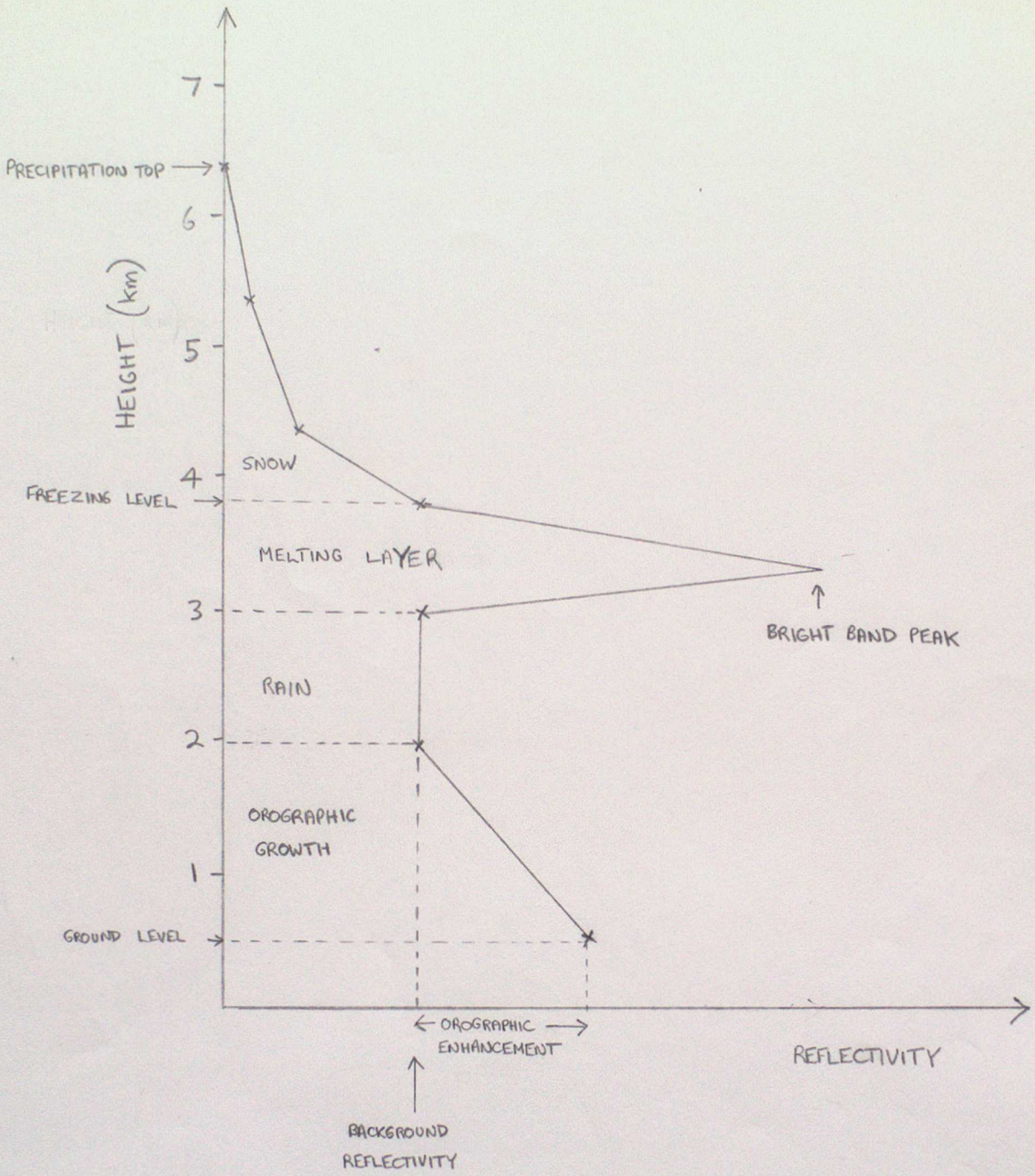


Fig 1 AN EXAMPLE OF AN IDEALIZED REFLECTIVITY PROFILE

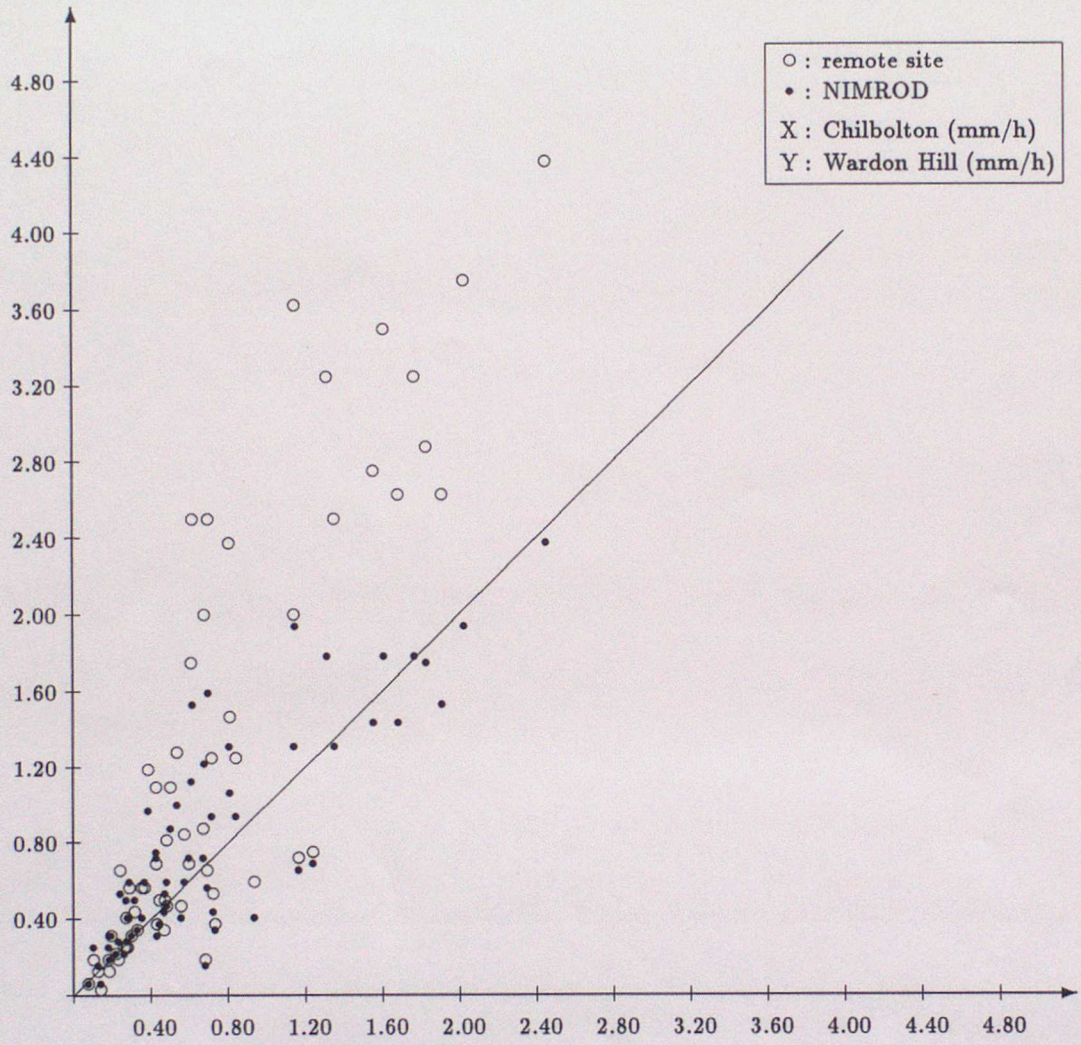


Figure 1: 21 Feb 1991 1500-1800UTC

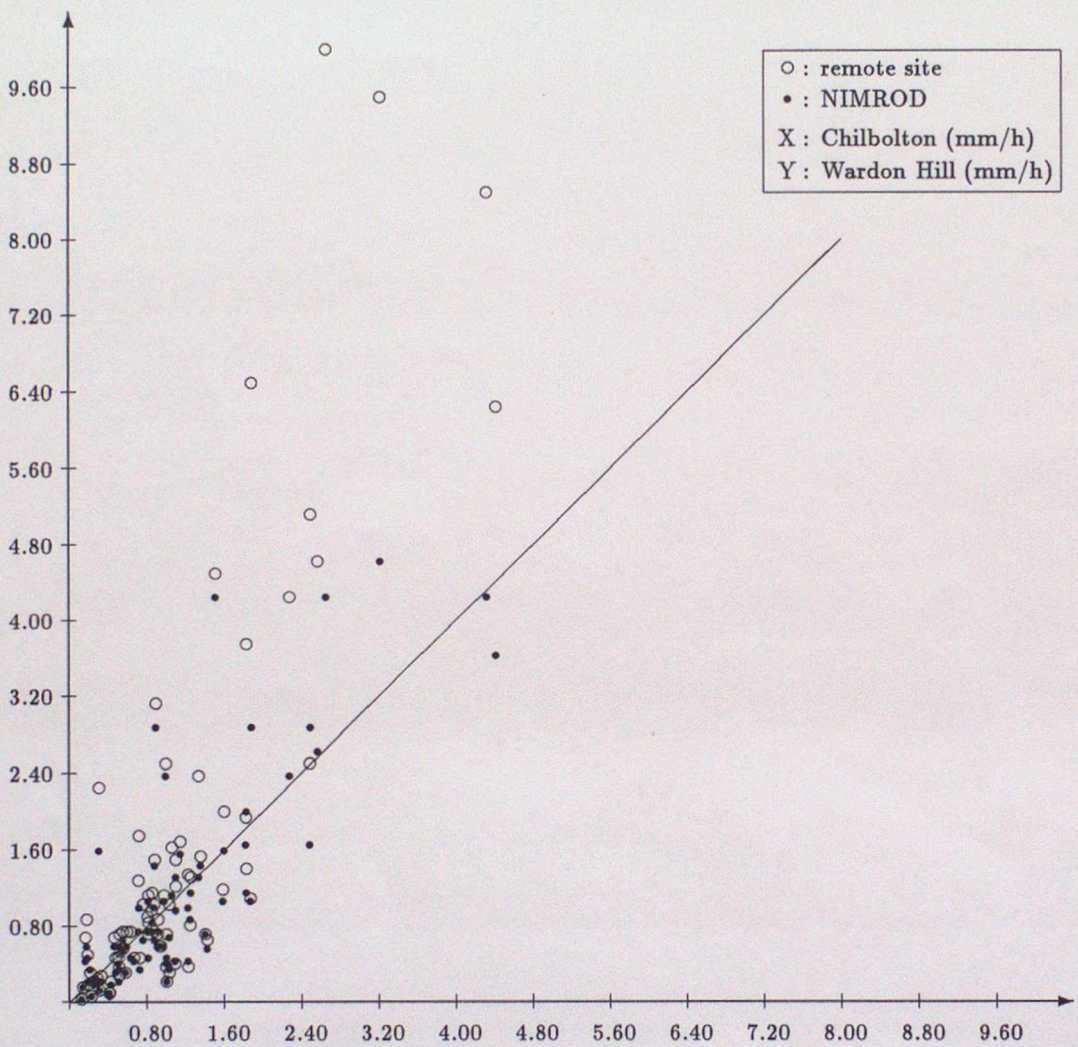


Figure 3 : 16 Mar 1991 1100-1530UTC

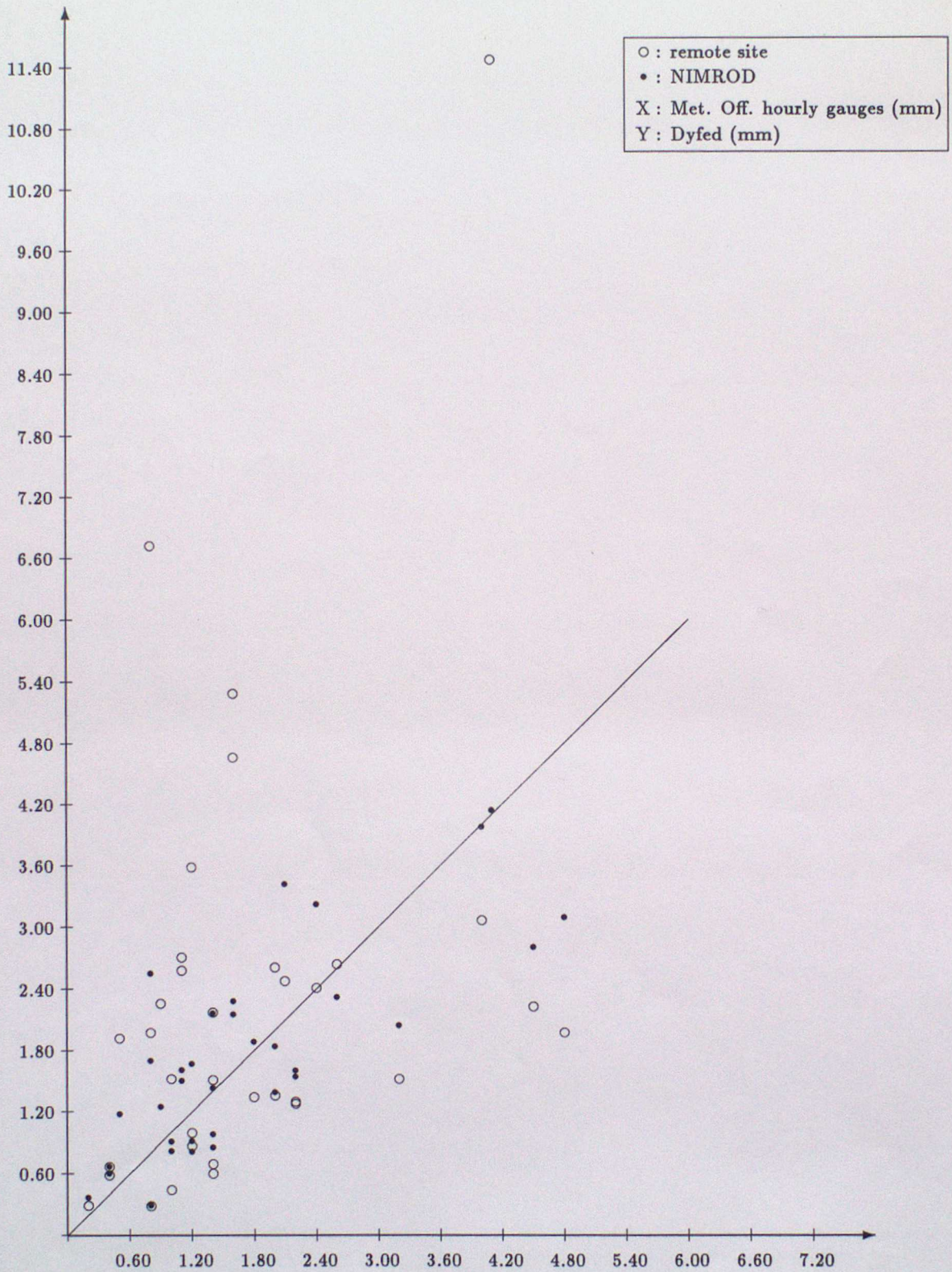


Figure 4: 9 Jan 1992 0000-0400UTC zero fields excluded

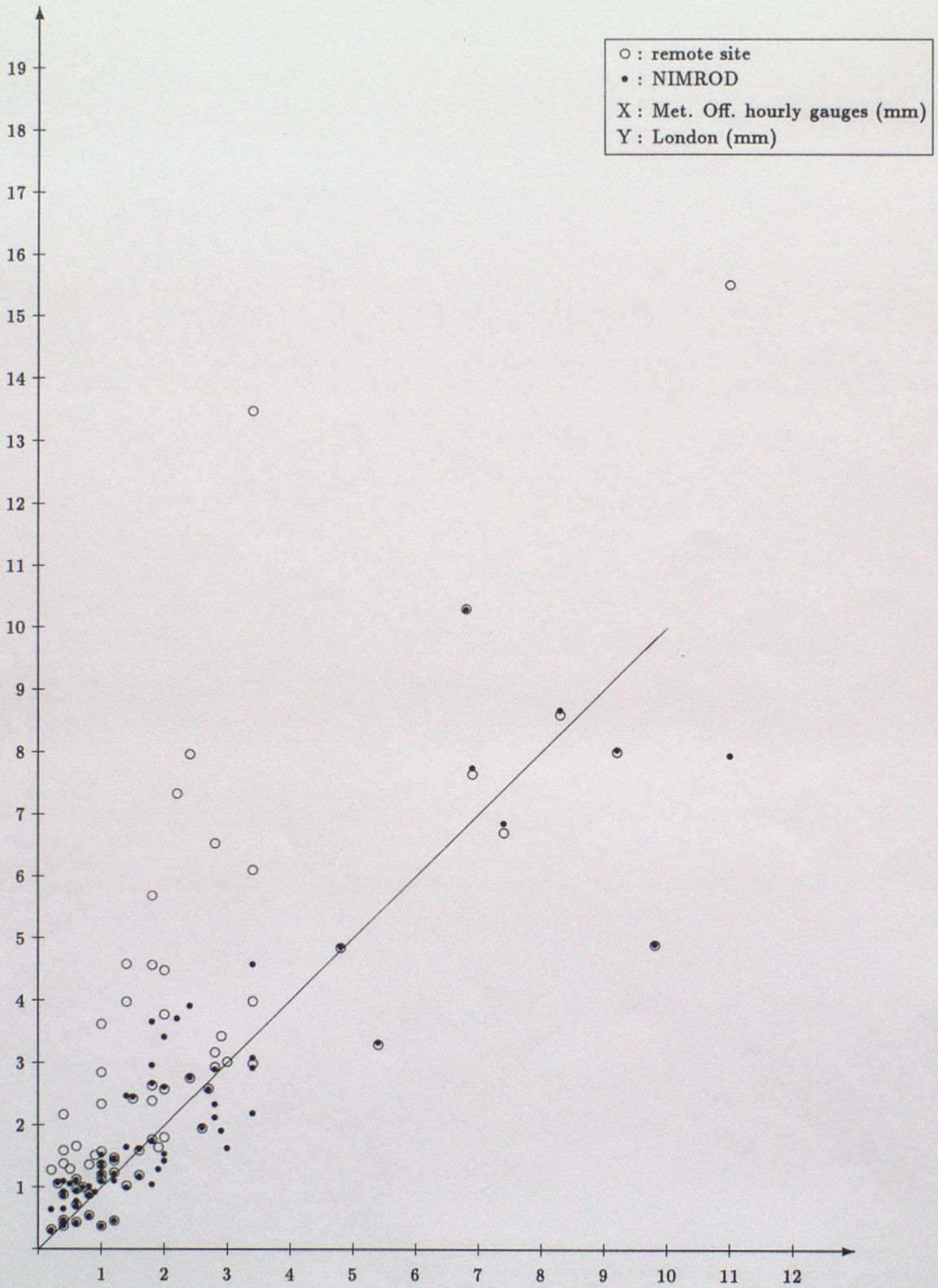


Figure 5: 3 Jul 1992 1200-1800UTC zero fields excluded

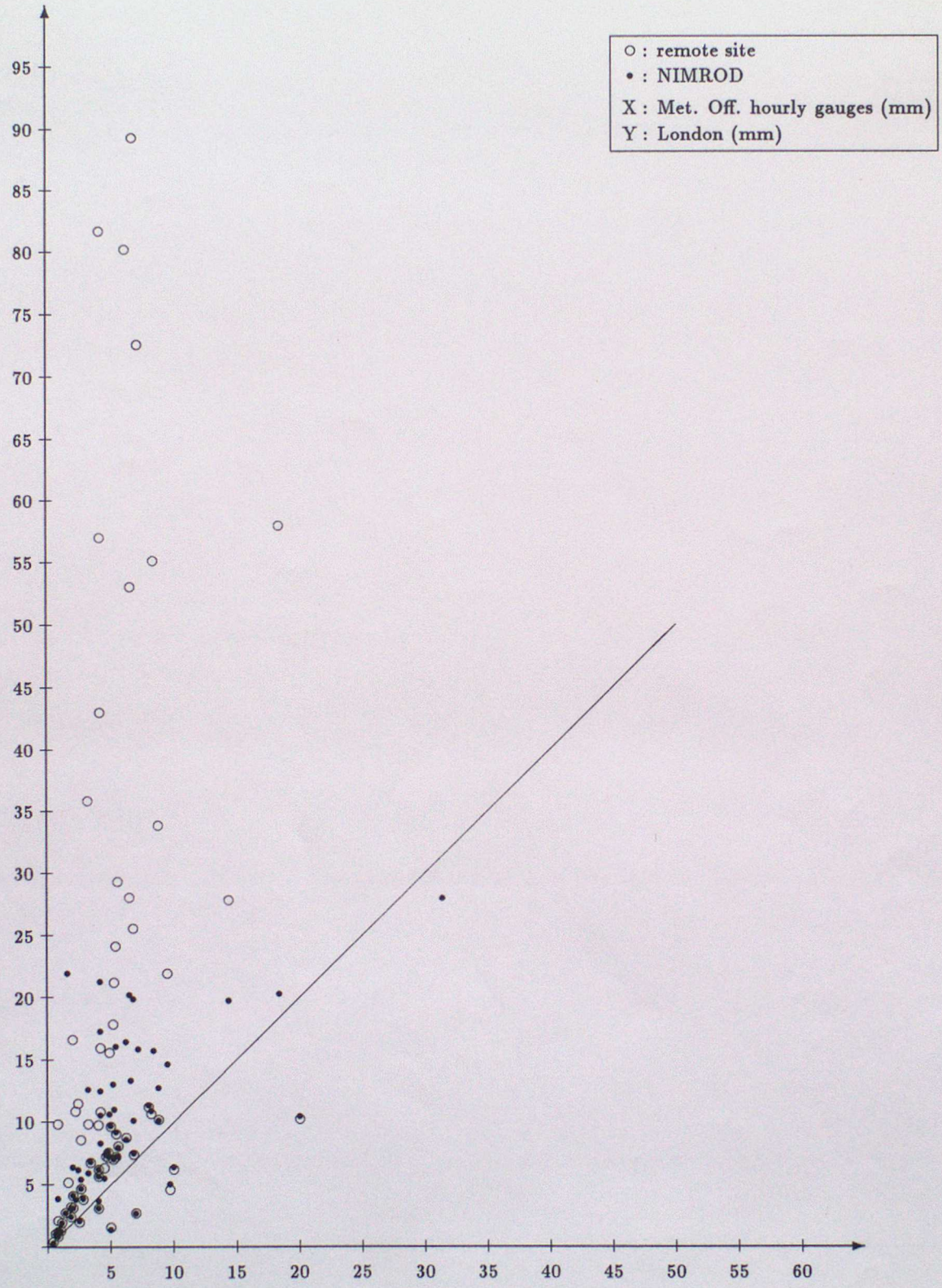


Figure 6 : 20 Jul 1992 1900UTC - 21 Jul 1992 0400UTC zero fields excluded